VALLEY WINDS

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1. INTRODUCTION

Valley scales range from 10^3 m to 10^6 m for length and from 50 m, say, to 5000 m for depth. In this review we shall not try to cover the whole range of scales but concentrate on valleys with lenghts $L \gtrsim 100$ km and depths $H \sim 1000$ m. This choice is partly motivated by the main concern of this seminar. Although the flow even in such "long" valley cannot be simulated in forecast models it is likely that the circulation in these valleys interacts with the flow resolved in a limited area fine-mesh model. Moreover, population density tends to be large in such valleys and the valley wind regimes have an impact on the air quality in cities like Innsbruck (Vergeiner et al. 1978). Valleys of this scale exist all over the world. Here we shall mainly be concerned with the large Alpine valleys like the Inn valley, or the Rhone valley. There are reasons for this bias: long research traditions and relatively good coverage by synoptic and climatic data.

The flow in these valleys is induced by solar heating and by the flow over the valley linked to the synoptic situation. We shall exclude here the latter effect and restrict our attention to the thermally driven valley flows with their pronounced daily cycles. Summaries of the observational aspects of thermally driven valley winds can be found in Wagner (1938), Defant (1951), Egger (1983) and Vergeiner and Dreiseitl (1986). Most of the material presented here is taken from the latter reference. A review of theoretical work on thermally driven flow in long valleys has not been written yet, partly because there is little to write about. Intensive efforts have been made to simulate the flow in relatively short valleys (see Pielke, 1985, for a review) but there are only few theoretical papers on the flow in long valleys. This is not surprising. The topography of the Inn valley, for example, is extremely complicated (Fig. 1). The side valleys and tributaries of the main valley form an intricate pattern. A detailed explicit simulation of the thermally driven flows in such a terrain is clearly impossible. At the moment modeling can at best aim at an understanding of the gross features of the observed flows.

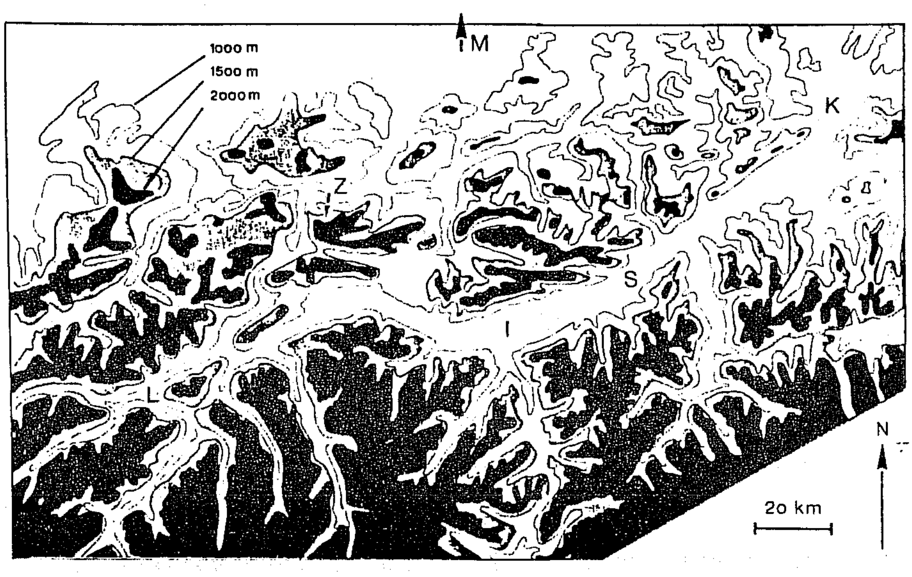

Fig. 1 Relief of Tyrol with the Inn valley as the main valley. Z = Zugspitze; M = 80 km to Munich; L = Landeck; I = Innsbruck; S = Schwaz and K = Kufstein. (From Vergeiner and Dreiseitl, 1986).

2. Observations.

Defant's famous scheme (1951) summarizes the daily course of slope and valley winds as observed. Upslope winds set in almost instantly when the sun begins to heat the slopes. The up-valley winds, however, start several hours later. According to Dreiseitl et al. (1980) up-valley winds in Innsbruck set in by about 9 a.m. in summer and as late as 1 p.m. in winter (Fig. 2).

Typical wind speeds are $\sim 5\ ms^{-1}$. In the evening the wind reverses and one observes down-valley winds throughout the night.

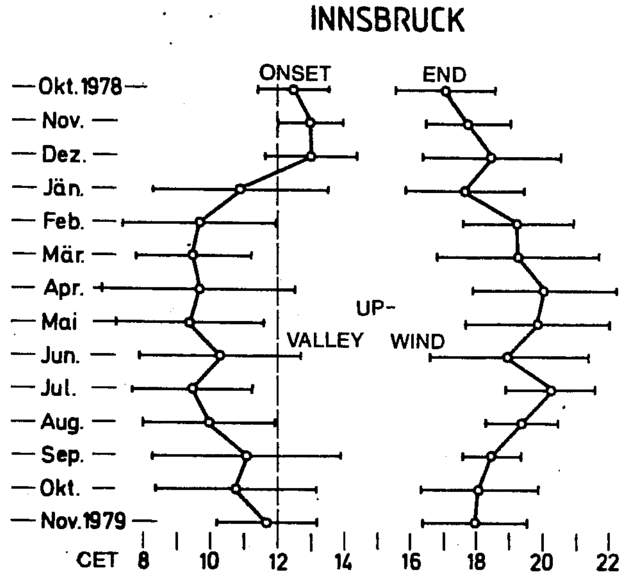

Fig. 2 Average monthly onset and end of up-valley wind phase at Innsbruck. After Dreiseitl et al. 1980.

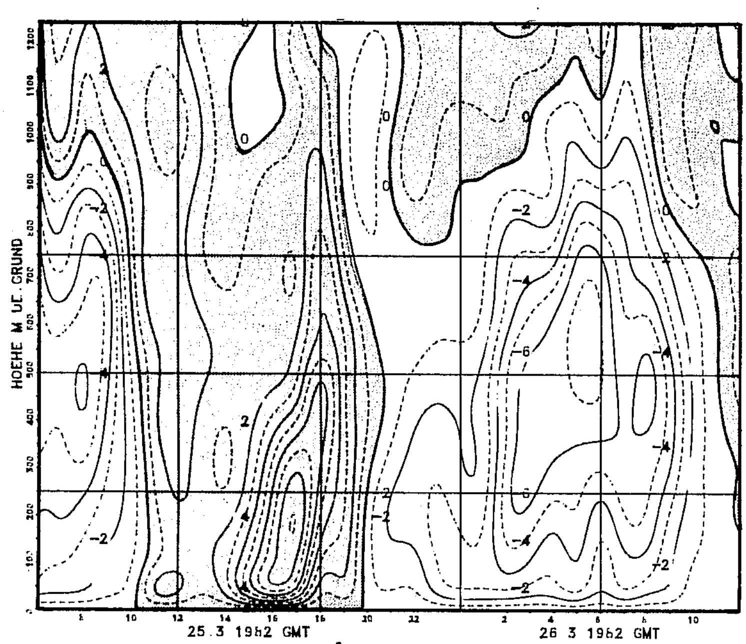

Fig. 3 Along-valley wind (ms^{-1}) at Niederbreitenbach in the Innvalley at two consecutive days. Stippled: up-valley wind. The crest level is about 1100 m above the valley bottom. From Freytag, (1986).

In Fig. 3 we show the valley winds as observed during the field experiment MERKUR in the Inn valley. On that occasion the down-valley flow was more intense than the up-valley flow. Note the abrupt changes of wind direction throughout the depth of the valley.

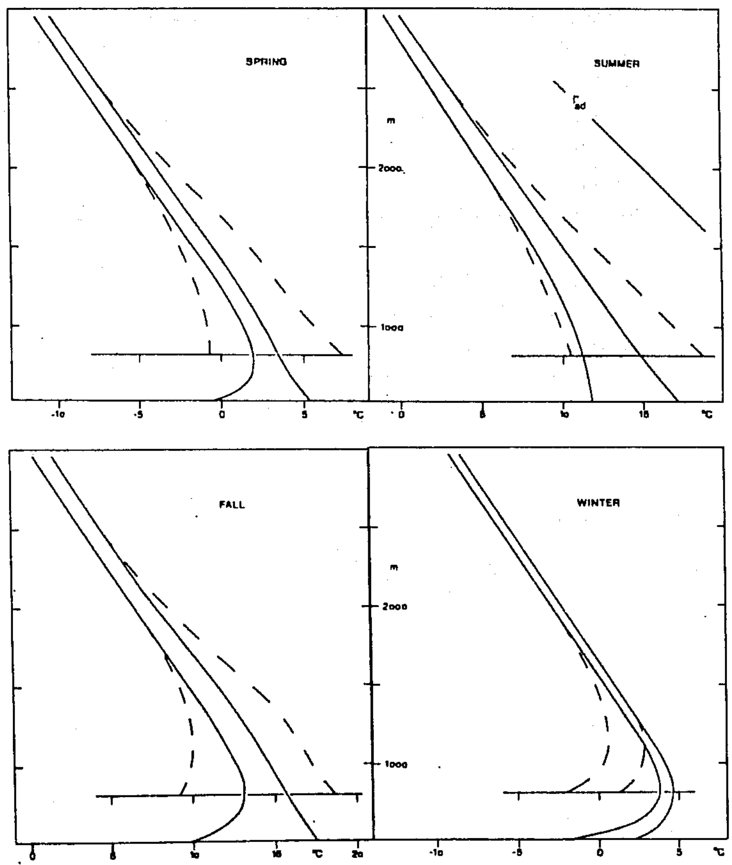

Fig. 4 Schematic profiles of the temperature for Landeck (821 m a.s.l.; dashed) and Munich (529 m, solid) at 06 Z and 15 Z for all seasons (after Nickus and Vergeiner, 1984).

The winds in the Inn valley appear to be linked to a circulation of Alpine scale. During the day there is low-level flow towards the Alps over the foreland (Wagner, 1937). As for the thermal state of the valley atmosphere it is well known that the amplitude of the daily variation of the temperature is larger in valleys than over the plain. In Fig. 4 we show temperature profiles for Landeck (see Fig. 1) in the Inn valley as compared to Munich in the Bavarian foreland for 06 Z and 15 Z, approximately the times of minimum and maximum temperature. Over the plain the daily range of the temperature decreases quickly with height. This is not so in the valley. Moreover, maximum (minimum) temperatures in the valley are higher (lower) by almost 3 degrees Celsius.

These differences in temperature can be converted to differences in pressure using the hydrostatic equation provided we know the pressure at crest height. Vergeiner and Dreiseitl (1986) point out that the mountains-foreland pressure difference is small at crest height. Correspondingly we have to expect considerable pressure differences at low levels linked to the differences in temperature.

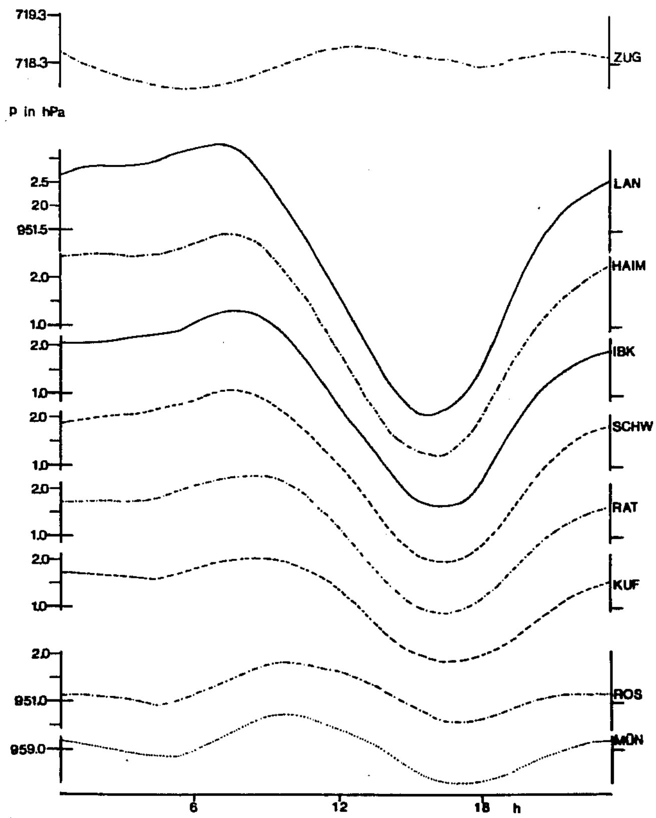

Fig. 5 Diurnal variation of pressure, averaged over six fair-weather days in September 1982 at six stations in the Inn valley (LAN = Landeck, 821 m; HAIM = Haiming, 692 m; IBK = Innsbruck, 579 m; SchW = Schwaz, 541 m; RAT = Rattenberg, 513 m; KUF = Kufstein, 508 m), for two stations in the foreland (ROS = Rosenheim, 445 m; MÜN = Munich, 529 m) and at Zugspitze. The six valley stations and Rosenheim are reduced to the height of Innsbruck. After Vergeiner and Dreiseitl, 1986.

In Fig. 5 we show the amplitude of the diurnal cycle of the surface pressure for several stations in the Inn valley and the foreland. It is seen that the amplitude of the pressure wave increase towards the valley's interior. Maxima and minima of the pressure occur more than an hour earlier in Landeck, the innermost station, than in Kufstein, at the valley mouth. The pressure difference between Landeck and Rosenheim is more than 3 hPa in the afternoon. While the mountain-plain pressure difference appears to be small above the Alpine crest height, this is not so for the Himalayas. During the day a low pressure system resides over the Tibetan plateau and high pressure is found at upper levels. During the night it is the reverse (Fig. 6).

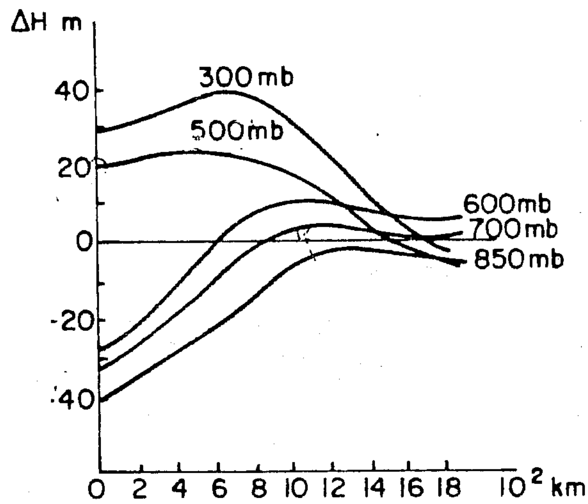

Fig. 6 Spatial distribution of height differences between 00 and 12 GMT at various pressure levels over Tibet (after Sang and Reiter, 1982). 0 corresponds with Lhasa.

Recent observations in the Vali Gandaki (Neininger, 1986; Reinhardt, 1986) showed that valley flows in valleys leading to the Tibetan plateau can be very intense. In Fig. 7 we show the winds at Jomosom, a little village half-way between the valley mouth and the Tibetan plateau. It is seen that the up-valley wind during the day is rather strong while the nocturnal outflow appears to be weak.

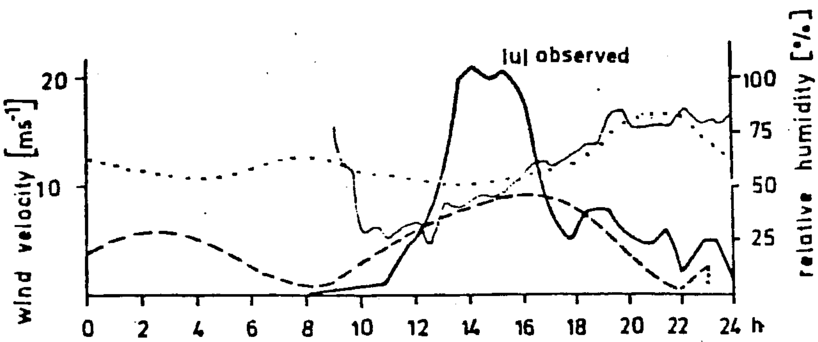

Fig. 7 Wind velocity (ms^{-1} and relative humidity on 2 February 1985 in Jomosom (bold). Data from Neininger (1986). Model results from Egger, (1987; u dashed; humidity dotted).

3. PHYSICAL CONCEPTS.

Why do we observe these diurnal valley-plain gradients of temperature and pressure? There appear to be two competing effects. First there is the "volume effect" as quantified by Steinacker (1984). According to this concept it is the difference of the volumes of air available to the heating in the valley and over the plain which is responsible for the temperature differences. Steinacker pointed out that the volume of air which must be considered when dealing with the energy budget of the valley atmosphere cannot be simply derived from a cross-section of the Inn valley. It is necessary to include all side valleys the circulation of which is linked to that of the Inn valley. If that is done one finds that more than half of the space available below crest height is filled by the mountains so that the corresponding ratios of volumes is $\gtrsim 2$. If we assume that the heating is essentially restricted to the air below crest height, say, this consideration immediately explains why the mean temperature of the valley atmosphere must be higher during the day than over the plain. We have, however, to add a mixing hypothesis in order to explain why the temperature in Innsbruck, say, should be higher than in Munich. To illustrate this point we show in Fig. 8 different columns of air with the same of unit area at the bottom. Column 1 is shallow and situated above a little plateau. Column 2 is deep and reaches from the valley-floor to the inversion height h_i. Column 3 is situated in the foreland. We assume that the same amount of thermal enery is available to these columns. Then it is obvious that the temperature of the shallow column will rise more strongly than that of the deep ones. It is also obvious that the mean temperature of the column 1 is larger than the corresponding mean temperature for an air volume in the plain where the heating is also restriced to layers with height $\lesssim h_i$. However, since the slope of the Inn valley bottom is rather small, there would be almost no valley-plain pressure gradient along the valley axis. To explain this observed feature we have to argue that the slope winds and the side valley circulations transfer part of the thermal energy received at and in the side valleys to the main valley (see Fig. 8).

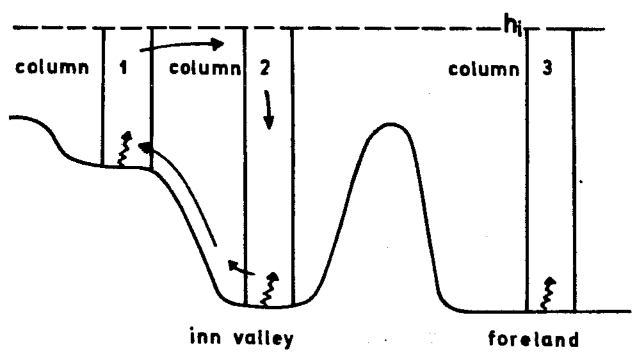

Fig. 8 Schematic of the heating of air columns in the mountains, the valley and the foreland. The arrows symbolize the slope wind circulation. h_i is an inversion height.

Thus the main valley is heated up as compared to the foreland and it is only then that a valley-plain pressure gradient can be established. A fairly detailed picture of this process emerged through the observational studies of Whiteman (1982), Brehm and Freytag (1982) and others.

There is, however, another possibility which may be important for valley flows in very large mountain massifs like the Himalayas (Fig. 9). Consider a valley cut out of a plateau and assume again that the heating is restricted to layers below a certain height h_i.

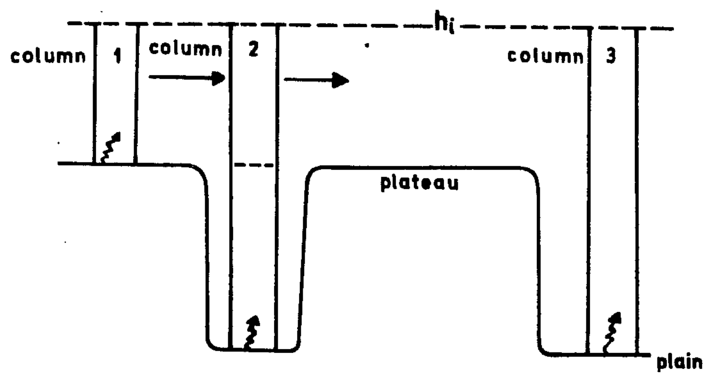

Fig. 9 Schematic of the plateau effect. The upper part of column 2 is replaced by warm air from the above the plateau.

Then the air above the plateau is much warmer than the air above the plain at the same height. The corresponding pressure difference will drive a thermal circulation even if there is no valley. We may assume, however, that the air above the plateau is advected towards the valley by a mean wind. The upper cool part of column 2 is replaced by warm air and, therefore, we obtain relatively low surface pressure at the bottom of the valley and we expect to find valley winds even if the volume effect is not important. Let us call this phenomenon "plateau effect". Both the volume effect and the plateau effect are closely related, of course. It may be worthwhile, however, to keep the distinction. For the "plateau effect" to be important there must a considerable mountain-plain pressure gradient well above the crest height of the mountain. This appears to be the case with the Himalays but not with the Alps. If the volume effect is operating, pressure gradients at the crest level may be weak. Nevertheless valley winds can be intense.

4. MODELS

It appears that no paper has been published yet on the diurnal valley-plain circulation for valleys like the Inn valley. What has been published is modeling work on the warming of the main valley by slope winds (e.g. Bader and McKee, 1985) and on the circulation in relatively short valleys (Pielke, 1985). As has been pointed out above, the warming of the valley atmosphere through the slope wind circulation forms an important part of the volume effect. So far this problem has been treated as two-dimensional and high-resolution simulations by Bader and McKee (1985) gave realistic heating rates for the valley atmosphere. Quite recently Brehm (1986) demonstrated that this heating effect can be simulated by a low-resolution model where just the equations for the total fluxes of mass and momentum in the slope wind layer and in the valley atmosphere are integrated. In Fig. 10 we show the changes of the temperature in the Eagle valley as observed and as given by the model of Brehm.

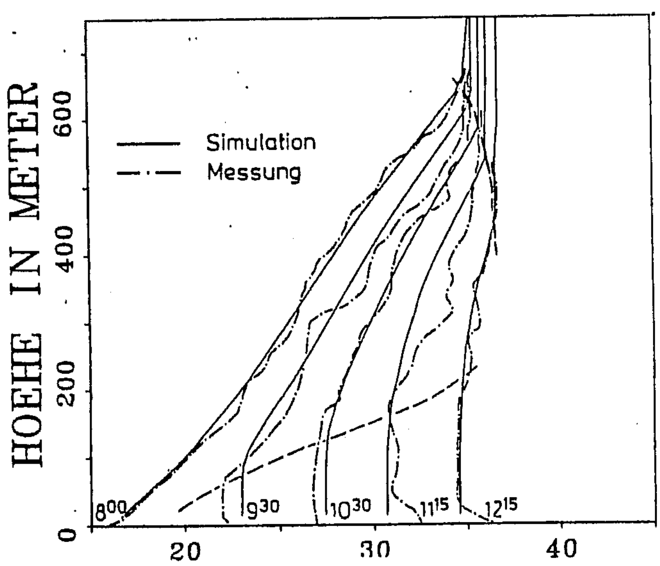

Fig. 10 Potential temperature simulated (bold) in the Eagle-valley on 16 October 1977 and as observed (dash-dotted) Dashed: boundaries of stably stratified domain. After Brehm 1986.

Obviously the model is capable of capturing the main features of the observed heating of the valley atmosphere. This suggests that simple models like that of Brehm can be used in three-dimensional calculations where the cross-valley circulation would be computed according to Brehm's scheme whereas the along-valley flow would have to be resolved in a standard grid. Such computations have not yet been carried out. If we turn to modeling of valley-plain circulations all the models proposed so far (Egger, 1986 a, b; Vergeiner, 1986; Egger, 1986 c) use highly simplified grid structures where the slope wind layers are not resolved at all. Vergeiner (1986) considers a flow domain which is subdivided into three sections: the foreland, a valley "tube" and a "basin" thought to represent the side valleys and tributaries of the main valley. The volume effect is prescribed as an enhanced heating in the basin. A balance of heating and momentum is computed for each domain and the horizontal fluxes through the boundaries of each domain are taken into account as well. However, the return flow aloft is not considered explicitly. Prescribing a diurnal heating cycle, Vergeiner was able to derive satisfactory valley winds with realistic intensities and lags.

A minimum resolution model has been proposed by Egger (1986 a) where the return flow aloft is also incorporated. This model is essentially a grid point model with four boxes (Fig. 11). Three of these boxes are of equal size but the fourth box represents the valley. The Boussinesq equations for shallow convection are written in finite difference form as appropriate for this grid. The heating of the model atmosphere is provided through a Newtonian heating term Θ_{oi}/T in the grid boxes at the ground. The temperatures Θ_{oi} can be interpreted as surface temperatures. The mixing time T characterizes the time it takes to heat the main valley via the volume effect. The heat is transferred upwards by eddy diffusion. The volume effect is invoked by prescribing higher surface temperatures in the valley than at the plateau. An energy balance is imposed for the heat supply:

$$\Theta_{o1} = \delta\,\Theta_{o2} + (1-\delta)\,\Theta_{o4} \qquad (4.1)$$

where Θ_{01} is the surface temperature in the foreland, Θ_{o2} that in the valley and Θ_{o4} that on the plateau. δ is the ratio of the valley width to the width of the domain. If $\Theta_{o2} > \Theta_{o4}$ in (4.1) we have additional heating in the valley and less heating at the plateau. This is in keeping with the concept of the "volume effect". Heat is transferred to the main valley from the surrounding mountainous terrain.

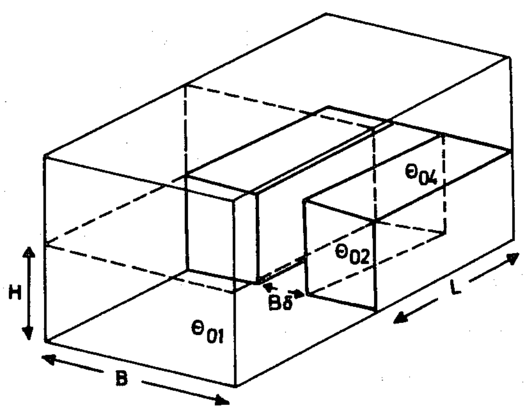

Fig. 11 Grid structure of a minimum resolution model for valley-plain flows. The model has four grid boxes. In particular, the valley is represented by one grid box of witdh B δ , length L and height H. All other boxes have the volume BxLxH. Also given are the surface temperatures at the respective surfaces.

If the diurnal heating cycle $\Theta_{oi} = \hat{\Theta}_{oi} \exp(i\omega t)$ with frequency $\omega = 2\pi \text{day}^{-1}$ is prescribed and if the model equations are linearized with respect to a stably stratified motionless reference state one finds for the wind u_1 at the valley mouth and for a very narrow valley ($\delta \sim 0$)

$$u_1 = \mathrm{Re}\left\{ \frac{\hat{\Theta}_{o2}\, H T \exp(i\omega t)}{L(\bar{\Theta}(T^2 H^2/(2T_B^2 L^2) + 2(2+i\omega T)^2)} \right\} \qquad (4.2)$$

(symbols: g gravity; H height of plateau; T mixing time $T = H^2/K$; K eddy diffusivity for vertical austausch; L length of valley; T_B buoyancy time of the mean state; $\tilde{\Theta}$ potential temperature of the reference state). According to Fig. 10, T is of the order of a few hours. In (4.2) it is assumed that there is no plateau effect. Thus (4.2) describes the valley wind induced by the volume effect. The interpretation of (4.2) is relatively easy. The winds in the valley are proportional to the excessive heating of the valley atmosphere through the volume effect, so that $u_1 \sim \hat{\Theta}_{O2}$. However, the valley wind lags the heating. This lag can be determined from (4.2) by looking at the real and imaginary part of the denominator in (4.1). One finds a lag of the order of about two hours for parameters typical of the Inn valley. Thus we would have to expect maximum inflow towards the valley at 2 p.m. The valley wind vanishes for very long valleys ($u_1 \rightarrow 0$ for $L \rightarrow \infty$). The foreland-valley temperature difference cannot be larger than Θ_{O2}. The corresponding foreland-valley pressure gradient is $\sim \Theta_{O2}/L$ so that $u_1 \rightarrow 0$ for $L \rightarrow \infty$, if Θ_{O2} is kept constant. An increase of T reduces the valley wind and a decrease of stability (increase of T_B) leads to an intensification of the valley winds.

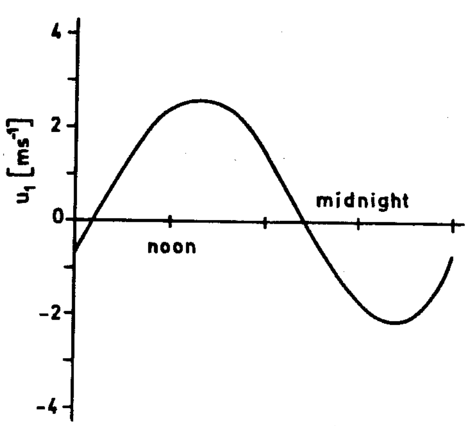

Fig. 12 Valley flow u_1 (ms^{-1}; up-valley flow for $u_1 > 0$) as obtained in the minimum-resolution model; T = 2 h. (After Egger, 1986 a).

In Fig. 12 we show u_1 as obtained in a nonlinear run with the four-box model where parameters have been chosen as thought to be typical of the Inn valley. The inflow lags the heating by about one hour. The maximum intensity of 2.6 ms^{-1} is reached shortly afternoon. The speeds thus obtained are relatively low but it must be kept in mind that u_1 is to be compared to the vertical average in Fig. 3. The inflow during the day is stronger than the outflow during the night. If the plateau effect is included we obtain an intensification of the wind intensity in the valley. There is, however, little evidence that the plateau effect is important in the Inn valley.

So far we considered simple box models of the valley-plain circulation with an extremely coarse horizontal resolution. Variations of the wind along the axis of the valley cannot be studied with such models. Let us now consider a one-dimensional model of the flow in the valley. Suppose we have a valley of infinite length and of depth H. The valley's mouth is at x = 0. We use a finite difference approximation in the vertical (Fig. 13).

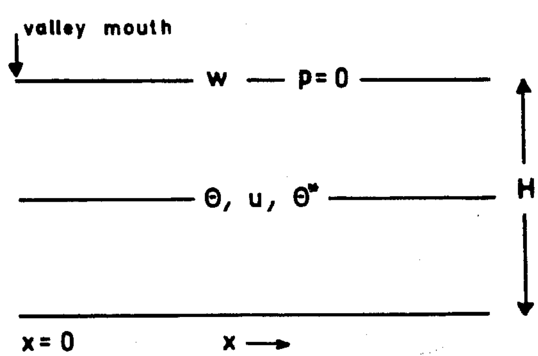

Fig. 13 Model structure underlying the flow equations (4.3)-(4.5).

The vertical velocity w is defined at z = H and the flow speed u is defined at z = hH/2 as is the deviation Θ of the potential temperature from that of the background state. The background state a is stably stratified with a buoyancy frequency assumed to be constant in time.

The first equation of motion reads at $z = H/2$

$$\frac{\partial u}{\partial t} = \alpha \frac{\partial \Theta}{\partial x} - u/T \tag{4.3}$$

where a linearization with respect to the motionless background state has been made and a Rayleigh damping is assumed. In addition the hydrostatic assumption has been made to compute the pressure gradient term in (4.3) with $\alpha = Hg/2\Theta$. Moreover it assumed that there is no pressure deviation on top of the valley in qualitative agreement with observations in the Inn valley (Nickus and Vergeiner, 1984). The first law gives

$$\frac{\partial \Theta}{\partial t} + \frac{w}{2} \frac{\partial \overline{\Theta}}{\partial z} = (\Theta^* - \Theta)/T \tag{4.4}$$

where Θ^* is the heating of the valley atmosphere due to the volume effect. We assume a diurnal variation of this effect so that $\Theta^* = \hat{\Theta}^* \exp(i\omega t)$ where $\hat{\Theta}^*$ does not depend on time and x.

The equation of continuity gives

$$\frac{\partial u}{\partial x} + \frac{w}{H} = 0 \tag{4.5}$$

We look for solutions

$$\begin{pmatrix} u \\ w \\ \Theta \end{pmatrix} = \begin{pmatrix} \hat{u} \\ \hat{w} \\ \hat{\Theta} \end{pmatrix} \exp(i\omega t) \tag{4.6}$$

Inserting (4.6) in (4.3) - (4.5) we obtain after simple manipulations

$$\frac{\partial^2 \hat{\Theta}}{\partial x^2} - \gamma^2 \hat{\Theta} = - \Theta^* (i\omega T + 1)/(T^2 N^2 H^2) \tag{4.7}$$

with

$$\gamma = 2(i\omega T + 1)/(TNH) \tag{4.8}$$

where N is the Brunt-Väisälä frequency of the reference state. The boundary conditions are $\hat{\Theta} = 0$ at the mouth and $\frac{\partial \hat{\Theta}}{\partial x} = 0$ for $x \rightarrow \infty$. Thus we assume that the temperature at the mouth is the same as the temperature in the foreland which is not influenced by the circulation in the valley. The solution to (4.7) is

$$\Theta = \mathrm{Re}\left[\hat{\Theta}^{*}(1-\exp(-\gamma x))\exp(i\omega t)/(i\omega T+1)\right] \quad (4.9)$$

This solution has several interesting features. For $|x| \gg |\gamma|^{-1}$ we obtain

$$\hat{\Theta} = \Theta^{*}/(i\omega T+1) \quad (4.10)$$

Off the mouth the temperature in the valley follows the heating due to the volume effect with a phase lag ε where $\tan\varepsilon = \omega T$. If we choose T = 2 h as before we obtain a lag of 1.5 hours for example. In (4.10) $\partial\hat{\Theta}/\partial x = 0$. Thus we do not a have a gradient of the temperature for $x > |\gamma|^{-1}$ and, therefore, there will be no wind in this part of the valley. Near the mouth one obtains a transition zone of width $D \sim TNH/2$ where the temperature adjusts to that in the foreland. With T = 2 hours, $N = 2 \times 10^{-3} s^{-1}$ for day-time conditions and $H \sim 1000$ m one obtains $D \sim 8$ km. Thus the transition zone at the mouth may be fairly narrow. This appears to be in agreement with observations (Freytag, 1985). At night where $N \sim 10^{-2} s^{-1}$ the transition zone may be wider. Since γ is complex the response near the mouth has a wavelike character. The wavelength of these waves is $2\pi D/\omega T$. The temperature changes propagate with the speed c = D/T into the valley.

According to (4.3) we have

$$u = \mathrm{Re}\left[\alpha\hat{\Theta}^{*}\gamma\exp(-\gamma x+i\omega t)/(i\omega T+1)^{2}\right] \quad (4.11)$$

The wind speeds are strongest in the transition zone near the mouth. The wind at the mouth follows the heating with a lag $2\omega T/(1-\omega^2T^2)$. For T = 2 hours the lag is 3.5 hours. It appears, therefore, that our simple model gives a reasonably good description of the diurnal variations of the valley winds near the mouth. However the model fails completely to explain the strong winds in the valley observed near Innsbruck or Landeck. What is observed in an increase of the amplitude of the diurnal variation of the temperature with distance to the mouth (see Fig. 5). It appears, therefore, that the volume effect becomes more pronounced if we move deeper into the valley. We can incorporate this feature into the model by prescribing

$$\hat{\Theta}^* = \hat{\Theta}_0^* + mx \tag{4.12}$$

where $\hat{\Theta}_0^*$ and m are constant. Let us restrict the flow to a valley of length L so that u (L) = 0, $\frac{\partial \hat{\Theta}}{\partial x}$ (L) = 0. Then the solution to (4.7) is for $|\gamma|^{-1} \ll L$:

$$\hat{\Theta} = [\hat{\Theta}_0^* + mx - \hat{\Theta}^* \exp(-\gamma x) - m\gamma^{-1} \exp(\gamma(x-L))] (1 + i\omega T)^{-1} \tag{4.13}$$

$$\hat{u} = \alpha T (m(1 - \exp(\gamma(x-L))) + \hat{\Theta}_0^* \gamma \exp(-\gamma x)] (1 + i\omega T)^{-2} \tag{4.14}$$

We rely on Vergeiner and Dreiseitl (1986) for a choice of the parameters in (4.13). These authors give the daily range of the valley mean temperature computed between Zugspitze summit and the respective valley stations over a fair-weather period in September (Table 1 of Vergeiner and Dreiseitl, 1986). We can adapt (4.13) to these observations by choosing $\hat{\Theta}_0^x \sim 0.75$ K and $m = 2.5 \times 10^{-5}$ km^{-1}. In Fig. 14 we show u(x) according to (4.14) at various times in a valley of 130 km length. The up-valley wind ($u > 0$) sets in at about 09 Z. At noon we can distinguish the transition zone at the mouth with a maximum speed of 6 ms^{-1}, the valley's interior where $u \sim 3$ ms^{-1} and the decrease of the flow speed near the valley head. Peak flow speeds occur at 15 Z.

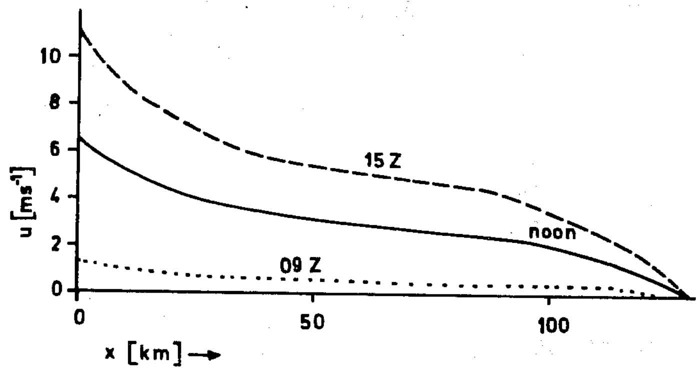

Fig. 14 Wind in the valley at various times according to (4.14).
$N = 0.002\ s^{-1}$, $T = 2$ hours, $\hat{\Theta}_o^x = 0.75$ K, $m = 2.5 \times 10^{-5} km^{-1}$.

Let us now turn to mountains of Himalayan scale with extremly long valleys. To produce valley winds for $L \sim 1000$ km, say, we have to invoke the plateau effect. If we were to rely on the volume effect we could induce valley winds in very long valleys only if we assume an increase of the intensity of the volume effect with distance to the mouth as has been done in (4.12). However, there are limits to this growth since the heat transfer from the side valleys to the main valley cannot increase without bounds. This means in turn that we cannot maintain a pressure gradient in a very long valley on the basis of the volume effect. We have to turn to models in order to see if the plateau effect is of sufficient strength to induce valley winds. In Fig. 15 we show the basic structure of a low resolution model of the diurnal circulation of a plateau with radial valleys.

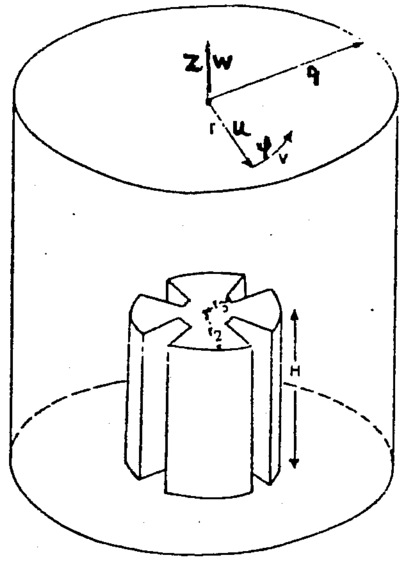

Fig. 15 Plateau with radial valleys in a circular domain. After Egger, 1986 c.

For reasons of symmetry it is sufficient to consider one sector of the circular plateau with just one valley leading into it. A minimum resolution model for this configuration has been developed which is based on six grid elements (Egger, 1976 c). The equations of motion on a f-plane, the first law and the continuity equations for dry air and for water vapor have been adapted to this grid structure. Numerical integrations of the model integrations have been carried out. The model atmosphere is heated from the ground. In the experiments to be discussed we exclude the volume effect, i.e. we prescribe the same diurnal march of the heating at the plateau and in the valley. In Fig. 16 we show the diurnal variation of the temperature above the plateau and above at the same height above the foreland.

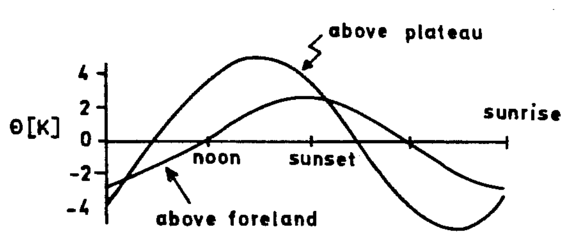

Fig. 16 Deviation Θ of potential temperature from reference state above the plateau and above the foreland at the same height. After Egger (1986 c).

The temperature above the plateau follows the heating with smaller time lag and a larger amplitude than above the foreland. Of course, this is due to the fact that the air above the plateau is closer to the source. This gradient of the temperature goes with a corresponding gradient of the pressure which drives a circulation above the plateau. However, the winds in the valley are much stronger and it can be seen (Fig. 17) that the circulation in the valleys contributes significantly to the radial mass flow of the plateau's diurnal circulation. The flow at the uppermost level is greatly reduced when there are no valleys. The circulation is modified by the influence of the earth's rotation. During the day one obtains a cyclonic flow around the plateau at and below the plateau and an anticyclonic circulation at the uppermost level.

The experiment has been repeated but with a volume effect prescribed. As had to be foreseen on the basis of (4.5) the effect on the valley flow was small. An attempt has been made to simulate the flow observations in the Vali Gandaki (Fig. 7) using this minimum resolution model. It is seen that the flow speeds in the model are of the same order of magnitude as those obtained. The time of maximum inflow is also captured reasonably well.

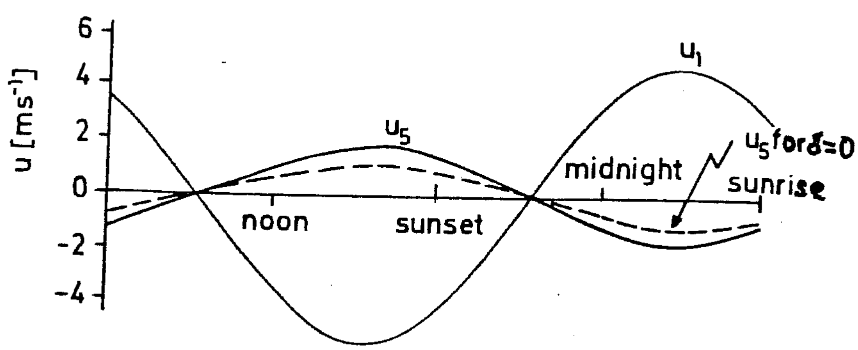

Fig. 17 Radial outflow u_1 in the valley (up-valley winds for $u_1 < 0$) and well above the plateau u_5. Dashed: u_5 for a situation without valleys, (from Egger, 1986 c).

5. CONCLUDING REMARKS

The problem of thermally driven flows in long valleys turned out to be extremely complicated. It is obvious by now that one cannot understand these flows without considering slope wind layers and the circulation in side valleys. So far, this volume effect has been parameterized rather crudely. Clearly more work is needed before we can claim to really understand the interplay of local circulations and the valley winds. The diurnal flows above crest height have little influence on the valley flows in the Alps although there must be some linkage for reasons of contiunity. It appears that the plateau effect becomes increasingly important when the scale of the massif and the length of the valleys increase. The model results suggest that the volume effect is of minor importance in the Himalayas. However, the treatment of both effects in the low resolution model is quite crude and it would be premature to draw firm conclusions on the basis of these model results. Moreover it is well known that the release of latent heat play a central role in the dynamics of the diurnal flow in the Himalayas. This effect is not incorporated in the models used so far.

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